2. CONCEPTUAL FLOW MODEL, PARAMETERS, AND CALIBRATION

The conceptual model of groundwater flow at Amchitka Island has several important components and assumptions, which are listed below and discussed in detail in subsequent sections.

- recharge and discharge characteristics governed by island hydraulics
- steady-state assumption
- isothermal conditions
- dual-porosity system where groundwater flow occurs principally through fractures
- homogeneity of hydraulic properties with vertical anisotropy
- limited impact of the nuclear tests on the island flow field

2.1 Model Components and Assumptions

Each of the model components is discussed below, followed by an evaluation of alternate conceptual models for groundwater flow.

2.1.1 Island Hydraulics

The fundamental setting of a freshwater lens overlying seawater in an island hydraulic system was described in the previous chapter. Though the basics of such systems were first identified near the turn of the 19th century (known as the Ghyben-Herzberg relationship), the complexity of the seawater intrusion problem has resulted in continued advances in the understanding of the processes into the 21st century. Despite these complexities, there are also assumptions that acceptance of an island hydraulic system allows in the conceptual model of Amchitka. Some of these assumptions have been corroborated by data collection on the island.

The first assumption is the existence of a freshwater lens beneath the island, sustained by recharge at the land surface. The presence of freshwater has been confirmed by drilling and sampling throughout the island, and the very presence of that lens requires recharge to be taking place. The large amount of runoff observed, numerous lakes, low permeabilities of the aquifer materials, and hydraulic heads near land surface all indicate that actual recharge volumes are low at Amchitka (discussed in more detail in the Data section), but recharge must be occurring nonetheless.

Following from the recharge assumption is the assumption that discharge occurs offshore, along the ocean floor. The presence of this discharge is inferred, and would be difficult to impossible to observe in the high energy environment off the coast of Amchitka. Sustaining the freshwater lens requires active groundwater circulation, however, so it is assumed that discharge does occur. An assumption of steady state (discussed next) requires that the volume of discharge will equal that of the recharge.

With recharge across the entire land surface and discharge in the adjoining sea, a groundwater divide is established in approximately the middle of the island, with flow directed to the sea on either side. Amchitka is elongated, such that the divide is assumed to run along its length, dividing groundwater flow to the Pacific Ocean on the southwest, and to the Bering Sea on the northeast. The position of the subsurface divide is assumed to coincide with that of the surface water divide, but even the surface water divide is difficult to discern in the Lowland Plateau area of the tests, as there is minimal topography and numerous lakes. The sensitivity of the groundwater divide assumption is therefore tested in the modeling.

The flow system resulting from the above assumptions is one of predominantly downward flow in the freshwater lens near the island center, arcing outward and finally upward to discharge beneath the ocean. Slower and smaller flux circulation in the seawater lens mirrors the freshwater cycle above. The downward portion of the freshwater flowpath has been confirmed for Amchitka through the measurement of hydraulic head (which decreases with increasing depth) and analysis of temperature logs (which indicate downward movement of cooler water). The upward directed reach of the freshwater flow system has not been measured due to the location of the wells nearer to the island center than to the coast. The saltwater circulation portion has also not been measured at Amchitka.

With flow directed perpendicular from the groundwater divide out to the sea floor, the island hydraulic problem lends itself to analysis in two dimensions. There is no driving force to create gradients along the island axis; flow outside of a plane perpendicular to the axis will only result from heterogeneities in the flow field. The consequence of the two-dimensional simplification, then, is one of underestimating spreading (dispersion) of contaminant plumes simulated in two dimensions.

2.1.2 Steady-state Assumption

The complete absence of pumping on Amchitka removes groundwater development as a cause of transient hydrologic responses. Rather, the question of the applicability of steady state conditions to Amchitka arises from temporal variations in natural processes. The processes identified as causing transient effects are tidal fluctuations, short-term variability in recharge, and long-term changes in climate that could affect sea level and possibly recharge rate. Transient responses caused by the nuclear tests themselves are addressed in Section 2.1.6.

With barometric correction, a close correlation was identified between tidal fluctuations at the shore of the Bering Sea and water-level fluctuations in UAe-1 (Fenske, 1972). The period, then, of the hydraulic head response is as rapid as that of the tide, and with a small amplitude (less than a tenth of a meter, plus and minus every day). The effect of this rapidly reversing transient pulse may serve to add dispersion to the contact between the freshwater lens and seawater, but neglecting it in the modeling will not affect the flow directions and velocities.

Short-term variations in recharge could occur due to weather variations from one year to another. Given the island's current condition at flow capacity, wetter years would only serve to increase runoff and thus not impact the groundwater system. A low precipitation year, or series of years, however, could possibly reduce recharge to the groundwater system. If such a reduction was sustained long enough, the depth to the transition zone would eventually be reduced, only to move downward again when average precipitation resumed. The major effect of such variations will again be to disperse the contact between the freshwater and saltwater.

Climate change involving glacial cycles affects mean sea level, which has a direct impact on the hydrology of an island. As mean sea level falls below current levels, the island's recharge area increases and the head for the saltwater system decreases. The net effect is to increase the depth of the freshwater lens. When sea level rises above current levels, the recharge area decreases, the head in the saltwater system increases, and the freshwater lens shrinks. Sea level varied by several tens of meters at Amchitka during the advances and retreats of Pleistocene glaciers (Gard, 1977). At least four major marine terraces have been mapped above the current island shoreline. The last major interglacial period caused a significant marine transgression at a level 37 to 49 m above present sea level. This transgression is dated at about 127,000 years before present. The last glacial advance ended about 10,000 years ago, at a sea level about 30 m below present. Sea level is believed to have been relatively stable at its present level for the past 2,000 to 4,000 years. Dudley *et al.* (1977) considered that time period long enough to likely allow for head adjustment, but question the response of water chemistry. Flushing of a relict deep freshwater lens will take much longer than head requilibration, especially given the slow velocities beneath the hydraulic transition zone. Dudley *et al.* (1977) suggest that freshwater circulation beneath Amchitka might have been as deep as 2,500 m during full glacial conditions, though the evidence of glaciers on the island itself calls into question the amount of recharge possible during the marine regressions.

Given the thousands of years that present sea level has been maintained, the steady-state assumption is considered reasonably valid for hydraulic head. As with the shorter-term processes discussed above, the impact of a glacial climate may principally be to have dispersed the transition zone, perhaps to the degree of leaving a relict, deep freshwater lens (as the last major sea level change was one of regression) that has yet to be displaced by the newer hydraulic regime. The different response times for the chemical and hydraulic systems to sea level changes are an important factor during the model calibration process, because the hydraulic data are considered more likely to represent equilibrium conditions. Given this impact, the relative response times are investigated in a sensitivity analysis.

2.1.3 Isothermal Conditions

Isothermal conditions are applied for the flow model, assuming that the flow system is dominated by the freshwater-seawater dynamics and that the effect of including geothermal heat would be relatively small. A significant factor in choosing this assumption is the large increase in computational load required by adding thermal conditions to the density flow solution, an already demanding series of computations as compared to a constant density problem. As discussed elsewhere, multiple realizations of the flow and transport are conducted to accommodate uncertainty. The merits of performing these Monte Carlo simulations were considered to outweigh the geothermal aspects. The error introduced through the isothermal assumption is investigated separately through a sensitivity analysis applying geothermal heat to the Milrow flow domain.

2.1.4 Dual-porosity System

Previous workers studying the hydrology of Amchitka Island have all concluded that groundwater flow occurs principally through fractures in the bedrock of the island. This conclusion has followed evaluation of geologic evidence gained through mapping and drilling, and analysis of hydraulic information from aquifer tests and cores. Basically, the very low permeability of the unfractured material, coupled with the presence of fractures and joints, leads to the conclusion that flow would be extremely limited other than in fractures.

Regarding the Banjo Point Formation investigated for Long Shot, the U.S. Army Corps of Engineers and USGS (1965) conclude:

> *The bulk of the rocks of the Banjo Point Formation have a porosity ranging between 4 and 26 percent but seem to have little interstitial permeability. The andesitic sills also have very little interstitial permeability. Rather, most of the permeability of these rock units seems to be in the fractures of the rocks and the permeability of these fractures seems to differ according to rock type and structural features.*

The U.S. Geological Survey also investigated the hydrology of the Long Shot area and concluded that the groundwater system consisted of interconnected fractures, based on observations made during drilling, examination of cores, and results of pumping tests (Gard and Hale, 1964). Gard and Hale note that though the rock has high porosity, the effective porosity was only in the fractures.

Fenske (1972) observed that lithologic descriptions, physical property measurements, and geophysical logs indicate that the island aquifer system consists of secondary, fracture-created porosity and hydraulic conductivity superimposed on primary intergranular porosity and hydraulic conductivity. He used fluctuations in the water level at UAe-1 in response to barometric and tidal fluctuations to evaluate the two-component flow system at Cannikin.

Deep exploratory holes, from 1,000 to 2,134 m in depth, were drilled and tested at Sites B through F. Pumping and slug-recovery tests in zones isolated with straddle packers were performed in the holes. Dudley *et al.* (1977) report that "Results of the hydraulic testing point strongly to fractures (joints and faults) as the primary avenues for fluid movement."

Based on these assessments and the underlying data, it is assumed here that the groundwater flow system at Amchitka is a dual-porosity system. The primary flow system is considered to be through fractured rock. This characterization is achieved principally through assuming very low effective porosity values, consistent with fractures rather than porous media. Interaction between the fracture flow system and the secondary system found in the high-porosity matrix blocks is allowed only through diffusion during transport (discussed in the conceptual transport model section). The fracture flow conception is considered to be a conservative assumption for the modeling. The use of very low effective porosity values directly relates to faster groundwater velocities. A porous medium assumption, with attendant higher porosity, would result in slower velocities and longer transport times.

2.1.5 Homogeneity of Hydraulic Properties with Vertical Anisotropy

Heterogeneity in hydraulic properties is the rule rather than the exception in subsurface geologic units. This heterogeneity, or spatial variability, results from small-scale to large-scale variations in geologic properties that in turn control water movement. Examples of features controlling hydrologic spatial variability are clay content in a formation, degree of fracturing, and depositional patterns. Though heterogeneity on some scale is always present, the need to explicitly include it in a numerical model of groundwater flow depends on a number of factors. The primary impact of heterogeneity on a groundwater transport problem is in representing the spreading of a contaminant plume resulting from the tortuous path that water takes due to variations in the flow field caused by the spatial structure of the aquifer. Including heterogeneity in a model increases the amount of data required (to describe the spatial structure), and causes reliance on parameters that are very difficult to determine, such as the correlation scale. The alternative is to consider the hydraulic properties homogeneous and treat the spreading process caused by heterogeneity through a macrodispersivity term in the flow equation (Gelhar and Axness, 1983; Hess *et al.*, 1992).

Identifying the preferred approach is based on the site data. The degree of variability in question needs to be evaluated. Then the availability of data to support a description of the spatial statistics of the hydrologic flow system must be considered.

Groundwater beneath the testing areas occurs in a variety of volcanic and sedimentary rock types, the vast majority of which is breccia, basalt, or a combination thereof. Analysis of hydraulic data from the extensive testing using straddle packers in island boreholes found a range of five orders of magnitude in hydraulic conductivity (K). As discussed in detail in the section on flow parameters, division of the K data into lithologic classes suggests that the mean K of the breccias may be lower than the mean K of the basalts. This difference is only "suggested," due to overlap in the data ranges and the small number of tests involved (only 13 in breccia intervals, 5 in basalt intervals, as the majority of packer seats straddled intervals containing combinations of lithologies).

It is important to note that the hydraulic testing is weighted toward the more permeable units, as these were intentionally singled out for packer isolation. Only about 10 percent of the section in the freshwater lens was considered permeable enough to hydrotest at Cannikin (Fenske, 1972). Zones that appeared to have extremely low K were not tested. Thus, though the range in hydraulic conductivity for the permeable portions of the breccias and basalts may overlap, these portions are interspersed through the subsurface with large sections having much lower K. It also follows that there are units of even higher K that may have been missed in the testing, or were included in a packed interval with units of lower K such that a lower averaged value was measured.

The close correlation between hydraulic head in UAe-1 and tidal fluctuations indicates that the aquifer responds with average characteristics rather than as extensive homogeneous layers of differing characteristics (Fenske, 1972). Thus, the layered heterogeneity behaves as an equivalent homogeneous unit. This response is also consistent with flow predominantly through fractures, as lithologic differences would be of lesser importance than structural features. However, Claassen (1978) identified four main zones of different horizontal hydraulic conductivity in the undisturbed subsurface system at Cannikin. These zones are not described or explained in his report, and analysis of the testing from UAe-1 and UA-1 does not suggest these zones. Claassen's study was focused at a smaller scale than the modeling presented here; he was examining flow through the Cannikin chimney and post-test hole during infilling that required considering small-scale variations in flow properties.

The small number of wells having hydraulic data, and large spacing between them, precludes spatial analysis at the scale of each site, though Cannikin does have one well between ground zero and the coast (HTH-1). The majority of hydraulic information is available for one borehole at each location. In terms of geologic information, four coreholes were drilled to investigate the subsurface geology at Long Shot. Three of these were drilled with a maximum separation of 183 m, with two directionally drilled (EH-3 and EH-6) to intersect the vertical hole (EH-5). A fourth hole (EH-1) was located 335 m to the north. Geologists concluded that the holes could only be correlated with each other in very general terms (U.S. Army Corps of Engineers and USGS, 1965). When the holes are within 30 m or less, several clear-cut correlations can be made. Working upward (and farther apart), several overlying gross units of one predominant lithology (mostly breccia, or mostly clastic) can be identified, but the precise top and bottom cannot be picked with certainty. Many smaller units, to 7.5 m thick, appear in one hole and nowhere else. No satisfactory correlations could be made across the 335 m to EH-1. Comparison with outcrop studies of the sea cliffs identified the "extremely irregular deposition of these breccias and fine clastics" (U.S. Army Corps of Engineers and USGS, 1965). It is noted that extreme lenticularity is the rule rather than the exception. The same lithologic types and variations were observed in both cross section and the core holes, with a lack of lateral continuity in both. At Cannikin, the lithology in UAe-1 and the emplacement hole UA-1 correlated well in terms of large-scale stratigraphic units, but the thickness of individual units was found to change drastically in the 90 m between holes, and some were found to pinch, swell, bifurcate, and even disappear. These lateral changes were not considered unusual for volcanic rocks (Lee and Gard, Jr., 1971).

Given the lack of major differences between the hydraulic conductivity of the rock types in the testing areas, coupled with a lack of ability to confidently project the locations of the units any distance from the boreholes, a homogeneous model was chosen for the numerical flow model. The heterogeneity known to be present is addressed in the model through the use of a macrodispersivity term. The heterogeneity assumption is consistent with the assumption of a dual-porosity flow system, where flow is predominantly through fractures that are distributed through all the lithologies present.

2.1.5.1 Vertical Anisotropy

Though the multitude of layers of varying lithology and varying hydraulic conductivity may be approximated as a single homogeneous unit, the layering results in that system being anisotropic. Flow perpendicular to the layers (vertically) occurs with more head loss than flow parallel to the layers (horizontally). Layered heterogeneity can lead to regional anisotropy values on the order of 100:1 (horizontal to vertical K), or larger (Freeze and Cherry, 1979). In fractured rocks, the anisotropy can be opposite, with vertical flow favored over horizontal. At Amchitka, the fracture system is overlain on a strongly layered lithologic section, so that the anisotropy is considered to favor flow in the horizontal direction, though not to the degree of a non-fractured sequence of heterogeneous layers. The anisotropy assumed here for Amchitka is 10:1.

Previous workers have also considered the island to behave anisotropically. Dudley *et al.* (1977) state that flow in the vertical direction is generally retarded throughout the island. Claassen (1978) studied hydraulic and water quality data from Cannikin and concluded there was a low vertical hydraulic conductivity, with little vertical flow indicated. Though Fenske (1972) contended that the hydrologic system at Cannikin at an elevation below -400 m was isotropic, he acknowledged that hydrotest and geological data required anisotropy on a local basis and he included it in a travel time model.

2.1.6 Limited Impact of the Nuclear Tests on the Island Flow Field

The underground nuclear tests at Amchitka are assumed to permanently affect the hydrologic environment in the immediate vicinity of each test, but not to affect the flow field at large. Each nuclear test created a cavity, which then collapsed and formed a rubble chimney above. Though there was surface collapse at Milrow and Cannikin, actual chimney formation did not propagate to the surface at any of the sites. Despite this, given the collapse and near-surface spalling, chimney-type properties are assumed to be continuous from the cavity to land surface at all three tests. These properties are increased vertical permeability and increased porosity.

Fracture intensity caused by an underground nuclear test varies with distance from the working point. General relationships are described by Borg *et al.* (1976) and are as follows: immediately adjacent to the cavity, and in the chimney, a zone of highly crushed rock is found, extending to a distance of about 1.3 cavity radii at the level of the test. A pervasively fractured zone then extends between 2.5 to 4 cavity radii. Beyond this is a region of widely spaced fractures with less frequent interconnection. Generally, at distances between 3.5 and 5.2 cavity radii, the compressive strength of the shock wave is too small to fracture the rock (the limit of shear failure). For many tests, the limit of shear failure coincides with the height of the chimney. Though tensile fracturing may take place beyond the shear failure limit, fractures are typically widely spaced and are considered to contribute little to an increase in overall permeability. Surface observations of nuclear testing effects focus on the impact of spalling. Spalling occurs when the compressional shock wave travels to land surface, causing surface layers to split away (and a temporary rise in land surface elevation), and subsequently slap down when the layers fall. This can result in fracturing of near-surface rock, confined to the upper tens to possibly a hundred meters below land surface, but unconnected to fractures from the cavity.

The International Atomic Energy Association (IAEA) modeled a variety of permeability configurations around a cavity and chimney and compared the results to data from two underground tests (IAEA, 1998). For the larger test (yield of 14.5 kt), the data were best matched by the models of no fracturing or fracturing only above the chimney. The smaller test (yield of 3.2 kt) was better simulated by a model of radially decreasing fracturing. This is consistent with observations of testing at the NTS, where larger tests are found to be better contained than smaller ones due to the establishment of a "stress cage" whereby the intense shock wave created by the test itself seals the near-cavity area from surrounding rock. Ultimately, the impact of local increases in K is limited by the hydraulics of the surrounding aquifer (*i.e.*, the water must eventually move into a less conductive area, and this controls the flux through the more conductive zone).

Nuclear tests also temporarily impact the flow field through transient pressure and temperature gradients. Immediately after the detonation, the vaporization of mass that creates the cavity also desaturates it. This causes hydraulic gradients to be directed toward infilling the cavity and chimney region prior to re-establishing the pre-existing regional gradient. Monitoring of this infill process at Cannikin indicates that it took less than two years to recover. This process could be incorporated into the modeling by preventing contaminant migration from the tests until after hydraulic re-equilibration. Radioactive decay would continue during this time, effectively decreasing the contaminant mass. Neglecting the infill process is therefore a conservative assumption applied here.

The transient temperature gradient is reduced during the infilling process, as groundwater cools the cavity; the thermal impact is greatest during hydraulic equilibration, when the cavity is infilling. Claassen (1978) presents a graph of the temperature history of the Cannikin cavity. A measurement 100 days after the test was about 200 degrees above the ambient pre-detonation temperature; this was reduced to about 25 degrees above ambient after 260 days. Based on vertical sampling in the post-test hole, Claassen concluded that "Large thermal gradients do not appear to have persisted in the Cannikin cavity. It is certainly possible that "hot spots" existed at time of abandonment of the site, but, if randomly distributed, they would contribute little to mixing of the cavity water." Additional examination of radiochemistry of vertically distributed samples in the Cannikin post-test well, led Claassen to observe that "Generalized cavity-wide convection of any magnitude is not apparent from examination of the available data." Based on this analysis, the transient temperature effects are not included in the bulk of the modeling, but are addressed through a sensitivity analysis.

2.1.7 Alternate Conceptual Flow Models

The conceptual flow model presented above is the basis of the bulk of the numerical modeling presented here. More complex, alternate, models can be imagined but are not supported by the data. In some cases, as noted above, specific alternative models are addressed through a sensitivity analysis.

For the island hydraulic system, the alternate model that could be pursued is in three dimensions rather than two. The basic island model is a two-dimensional problem, but there are variations that do require three dimensions to properly analyze. For the Amchitka testing areas, the impact of fault zones on the flow system, the impact of the chimneys created by the nuclear tests, and the impact of thermal processes, all require three dimensions. These alternate conceptual models are addressed in a sensitivity analysis, following the two-dimensional simulations.

An alternate conceptual model for the steady-state assumption would incorporate the temporal processes described above. Data on tidal fluctuations and long-term sea level changes exist; the magnitude and timing of short-term climate variations are unknown. None of these past processes are included in our modeling because their inclusion would not alter the current groundwater velocity fields (and travel times) calculated here. Forecasting future climate change is highly speculative and not attempted here. If another glacial climate occurred and sea level dropped below its present level, the depth of the freshwater lens would increase and travel times from Milrow and

Cannikin might decrease. Little impact would occur at Long Shot, as its cavity is already in the freshwater zone. If interglacial conditions of the magnitude indicated by the marine terraces occur, an increase in sea level will shrink the freshwater lens at Amchitka and travel times from all three cavities might increase. Such a change would potentially be most dramatic at Long Shot, if the transition zone rose above the cavity horizon.

Rather than a dual-porosity, fracture flow system, an alternate conceptual model is to consider groundwater flow at Amchitka as strictly in a porous medium. This conception was tested by assigning an effective porosity coincident with the core measurements at Amchitka, keeping the other hydraulic parameters constant. The result (discussed in section 4) is that travel times to the discharge area on the sea floor increase enormously (only 29 out of one hundred realizations showed breakthrough at the sea floor after 5.5 million years). Given this result, the alternate conceptual model of porous media flow is not pursued due to its lack of conservatism. Though not a question of conceptual model so much as a question of implementation of the conceptual model of fracture flow, an equivalent porous medium approximation of fracture flow is used here. This is driven by the scale of the hydraulic data, coincident with that of equivalent blocks rather than discrete fractures, and the many assumptions regarding fracture geometry that are required by a discrete fracture model.

The alternative to a model of homogeneity is a conceptual model of discrete aquifers and/or aquifers with spatially variable hydraulic properties. To implement such conceptual models with a numerical flow model would require numerous assumptions regarding spatial distribution of hydraulic units and properties that cannot be supported by site data. The result would be greater uncertainty in the modeling results due to the introduction of parameters with no supporting data. The disadvantage of the homogeneity assumption is that it neglects known system complexity. For the testing areas, the potential negative consequence of this would be overlooking fast transport pathways. Given the lack of correlation observed at Long Shot, it is doubtful that stratigraphic units would provide direct transport pathways from the test cavities to the ocean floor.

The U.S. Army Corps of Engineers and USGS (1965) present a conceptual model of groundwater flow for the Long Shot site that examines flow through an andesite sill. They consider several variations regarding the relative hydraulic conductivity of the andesite and surrounding units, but key to all the situations is an assumption that the andesite is continuous between the testing area and the discharge point at the sea floor. They conclude that fractures and their distribution are the most important features of the flow system, a conclusion also made by the U.S. Geological Survey, as their description of the Long Shot flow system does not single out the andesite (Gard and Hale, 1964). The possible impact of conductive features of large continuous length is evaluated in a sensitivity analysis regarding the possible effect of faults on the flow system. That analysis can also be applied to conductive stratigraphic features, though they are less likely than fractures in the Amchitka environment.

Another alternative to the assumption of homogeneity is presented by Dudley *et al.* (1977) in the form of a two compartment hydrologic system on the island. They identify a shallow, moderately permeable, groundwater system that readily accepts recharge and in turn discharges in springs and to streams and lakes. This is in contrast to the deep hydrologic system in the consolidated rocks, characterized by low permeability and very slow rates of groundwater movement. They believe that most groundwater beneath Amchitka moves in very local systems to discharge into lakes and streams, and that Amchitka is drained principally by streams that carry direct runoff and sustained base flow derived from discharge of the shallow groundwater system to the streams and lakes. Implementing this conceptual model into a numerical model will result in a dynamic upper hydrologic layer that is essentially de-coupled from the underlying flow system of interest. The upper boundary of the deep system would become a leakage boundary from the shallow system, a subtle difference from the recharge applied in a single-compartment model. The important aspect of the dual-compartment model, that the abundant precipitation at Amchitka is diverted from the deep groundwater system, is captured through a low recharge rate in the single-compartment model. There is no suggestion of an unsaturated zone between the shallow and deep systems (no perching), so that pressure heads can be expected to be continuous. Significant lateral flow in a shallow system might result in somewhat lower heads at the top of the model domain than simulated with the single-compartment model here. The net effect of including a discrete upper layer would be to add complexity and detail in a small portion of the model domain that is not of concern to the modeling objective, and would not result in more rapid contaminant transport than the model pursued here.

2.2 Flow Model Parameters and Supporting Data

2.2.1 Hydraulic Conductivity

Data from hydraulic tests in exploratory holes on Amchitka Island are presented and analyzed in a series of reports released by the USGS in the late 1960s and early 1970s (Ballance, 1970a, 1970b, 1972a, 1972b, 1972c, 1973b). Most of the tests involved isolating uncased or perforated cased intervals of the boreholes with inflatable straddle packers, and injecting or swabbing known quantities of water. Difficulties in obtaining firm packer seats in cased and uncased portions of some boreholes led to repeated tests in many intervals, and in several intervals the tests could not be completed without packer bypass, as noted by the USGS. The packer intervals ranged in length from 18.3 to 485 m, with an average length of 85 m. The USGS reports present values of relative specific capacity (RSC), rather than transmissivity or hydraulic conductivity, *K*, for most of the tested intervals. As opposed to specific capacity, RSC is derived from a short-duration test of a defined interval (packed interval). The RSC values reported in the USGS reports are generally estimated from injection data.

Unfortunately, the RSC data cannot be used directly in groundwater modeling efforts at Amchitka, which require a description of the hydraulic conductivity of the subsurface. Therefore, an independent analysis of the hydraulic data reported by the USGS was conducted. All of the time-recovery plots presented in the USGS reports were digitized and the data analyzed using the method of Cooper *et al.* (1967), resulting in estimates of *K* for 74 intervals in the six available wells. From these results, we selected the 42 values of *K* estimated from swabbing tests where no packer bypass was noted by the USGS (Table 2.1). Only swabbing test results were included because injection test results can be biased toward lower values of *K* if pores in the rock become plugged as water and suspended material move into the formation during the test. All values of *K* were calculated for standard temperature and salinity conditions owing to the lack of consistent reporting of salinity and temperature values in the tubing or packed intervals.

Table 2.1. Summary of hydraulic data from straddle packer tests on Amchitka Island.

Well Name	Depth (m below surface measurement point) Interval Top	Interval Bottom	Static Water Level*	K** (m/d)	$\log_{10} K$	Lithology***
Cannikin Site						
UA-1	1596.6	1647.5		6.0E-05	-4.2	Breccia
UA-1	1651.4	1702.9		7.0E-05	-4.2	Breccia
UA-1	1693.5	1753.8	35.0			
UA-1	1694.1	1769.4		1.2E-03	-2.9	Breccia
UA-1	1745.3	1805.6		3.1E-04	-3.5	
UA-1	1755.7	1808.1		7.6E-04	-3.1	Basalt
UA-1	1756.8	1968.4		1.9E-03	-2.7	
UA-1	1806.3	1859.9	36.4	2.6E-03	-2.6	
UA-1	1813.0	1873.3	38.7			
UA-1	1834.9	1895.3	37.7			
UA-1	1877	1914.5		1.9E-03	-2.7	
UA-1	1896.7	1968.4	36.4	2.8E-03	-2.6	
UA-1-HTH-1	80.5	125.0	3.9			
UA-1-HTH-1	128.3	179.5		6.5E-05	-4.2	Breccia
UA-1-HTH-1	183.5	234.7	4.0	1.1E-01	-1.0	Breccia
UA-1-HTH-1	227.4	278.6	5.7	8.2E-02	-1.1	
UAe-1	487.7	518.2	34.4	9.0E-02	-1.1	Basalt
UAe-1	542.6	563.9	34.4			
UAe-1	951.0	969.3	34.2	5.0E-02	-1.3	Basalt
UAe-1	1531.3	1580.1	33.1	1.1E-02	-1.9	
UAe-1	1645.9	1724.6	35.1	1.2E-02	-1.9	
UAe-1	1648.4	2133.6	33.9	2.4E-03	-2.6	
UAe-1	1724.0	1784.3	37.7	7.6E-03	-2.1	Breccia
UAe-1	1922.1	1982.4		4.0E-05	-4.4	Breccia
UAe-1	1966.0	2026.3		2.3E-03	-2.6	
UAe-1	2027.6	2133.6		1.2E-03	-2.9	
Milrow Site						
UAe-2	379.8	440.1	14.2			
UAe-2	933.9	994.3	27.0			
UAe-2	719.3	779.6		4.6E-02	-1.3	
UAe-2	994.2	1054.6		1.8E-04	-3.7	
UAe-2	1057.6	1127.8	19.7	1.3E-02	-1.9	
UAe-2	1057.6	1127.8	25.0			
UAe-2	1109.5	1169.8		1.8E-03	-2.7	
UAe-2	1164.3	1224.7		9.0E-04	-3.1	

Table 2.1. Summary of hydraulic data from straddle packer tests on Amchitka Island (continued).

Well Name	Depth (m below surface measurement point)			K^{**} (m/d)	$\log_{10} K$	Lithology***
	Interval Top	Interval Bottom	Static Water Level*			
UAe-2	1230.2	1290.5		9.0E-04	-3.1	
UAe-2	1292.4	1352.6		1.2E-03	-2.9	Breccia
UAe-2	1355.8	1416.1		1.5E-03	-2.8	
UAe-2	1521.6	1581.9		1.2E-03	-2.9	Breccia
UAe-2	1621.5	1681.9		4.6E-03	-2.3	
UAe-2	1725.8	1786.1		2.0E-04	-3.7	Breccia
UAe-2	1893.1	1980.6		8.0E-05	-4.1	
Long Shot Site						
EH-5	663.2	724.2	14.2			
EH-5	723.6	784.6	16			
EH-5	723.6	784.6	16.9			
Other Sites						
UAe-3	216.4	276.8	31.6			
UAe-3	318.8	357.2	32.1			
UAe-3	527.3	587.7	32.0			
UAe-3	1702.6	1747.7	115.0			
UAe-3	2032.4	2098.0		4.7E-04	-3.3	Breccia
UAe-6h	85.0	123.8	22.9			
UAe-6h	85.0	1493.5	32.7			
UAe-6h	85.0	2133.6	41.0			
UAe-6h	1074.1	1137.5		2.0E-03	-2.7	Basalt
UAe-6h	1236.9	1494.8		2.0E-03	-2.7	Basalt
UAe-6h	1498.7	1589.2	71.0			
UAe-6h	1503.3	1746.5	90.0			
UAe-6h	1560.6	1614.8		1.9E-03	-2.7	
UAe-6h	1617.9	1744.1		2.0E-03	-2.7	
UAe-6h	1702.0	1771.5		4.7E-03	-2.3	
UAe-6h	1774.6	1866.6		9.4E-03	-2.0	Breccia
UAe-6h	1906.8	2116.6		2.7E-03	-2.6	
UAe-6h	2019.6	2116.6		4.0E-04	-3.4	Breccia

Notes: *Measured static water levels from injection and swabbing tests having no packer bypass
**Analysis of USGS data from swabbing tests having no packer bypass
***Listed only for intervals that are entirely within a single lithologic unit and have an associated K value

The distribution of the $\log_{10}$-transformed K values is shown in Figure 2.1 and is notable for both its wide range of variability and for its overall relatively low values. Most of the Amchitka K values are between 1.0×10^{-4} and 1.0×10^{-1} m/d, a range that falls at the lower end of the range of K values reported for fractured rocks by Freeze and Cherry (1979) and volcanic rocks at the NTS by Rehfeldt *et al.* (1996).

Categorization of the K values into lithologic classes suggests that the mean K of the breccias may be lower than the mean K of the basalts (Figure 2.1). Note that the sum of the number of breccia

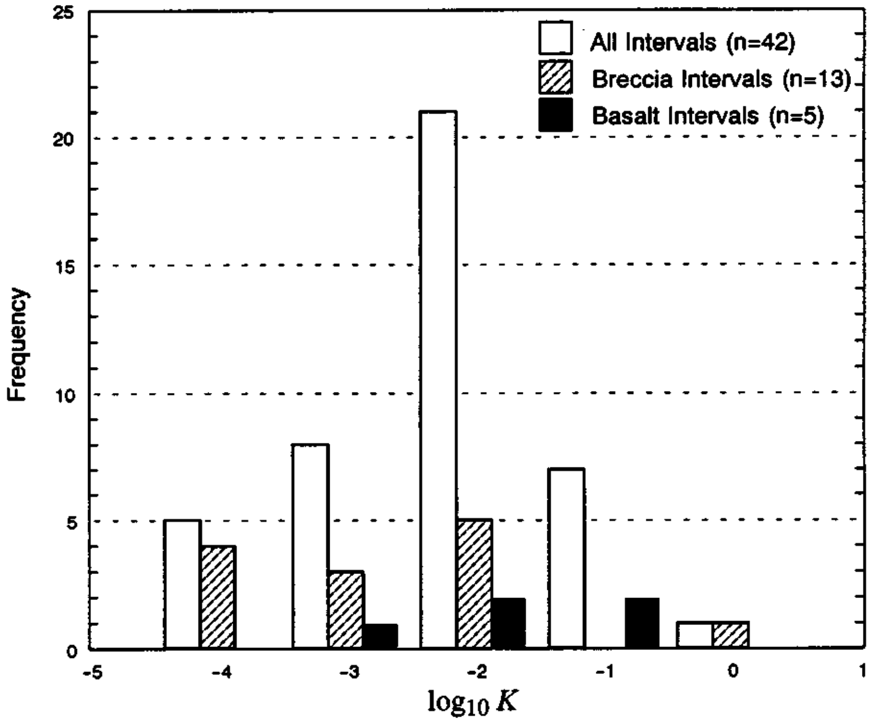

Figure 2.1. Distribution of $\log_{10}$-transformed K values estimated from straddle packer test data collected from Amchitka boreholes. Many intervals contained a mixture of breccia and basalt and could not be assigned to one lithology.

intervals and basalt intervals is less than the total because some intervals contained both, or neither, lithologic units. In any case, the spatial distribution of these lithologic units is very poorly understood, and owing to the lack of appropriate data, this heterogeneity is not directly incorporated in the flow model. The tidal fluctuations and water level responses in UAe-1 are not representative of layers with differing hydraulic characteristics, but rather of an averaged system (Fenske, 1972), supporting this approach. Some variation in K is evident between wells, although the overall ranges of the data far outweigh these differences (Figure 2.2). There appears to be a trend of declining K with increasing depth in two of the wells (UAe-1 and UAe-2), but this is not clear in all of them, due in part to the data distribution. The flow models do not include variation of K with depth, as no consistent quantifiable trends are evident in the data set.

2.2.2 Hydraulic Head

Water levels measured during hydraulic testing are used to represent hydraulic heads at the depths of the packed intervals (Table 2.1). Unfortunately, the effects of packer bypass and the incomplete recovery of water levels after swabbing in many tests reduced the number of reliable data available for characterizing the spatial distribution of hydraulic head; many of the values tabulated in the USGS reports are noted as "estimated" or "assumed" based on measurements made under similar conditions in other intervals or other boreholes. Of the data obtained from packer tests, only water levels that clearly indicate static conditions, either prior to an injection test or upon full

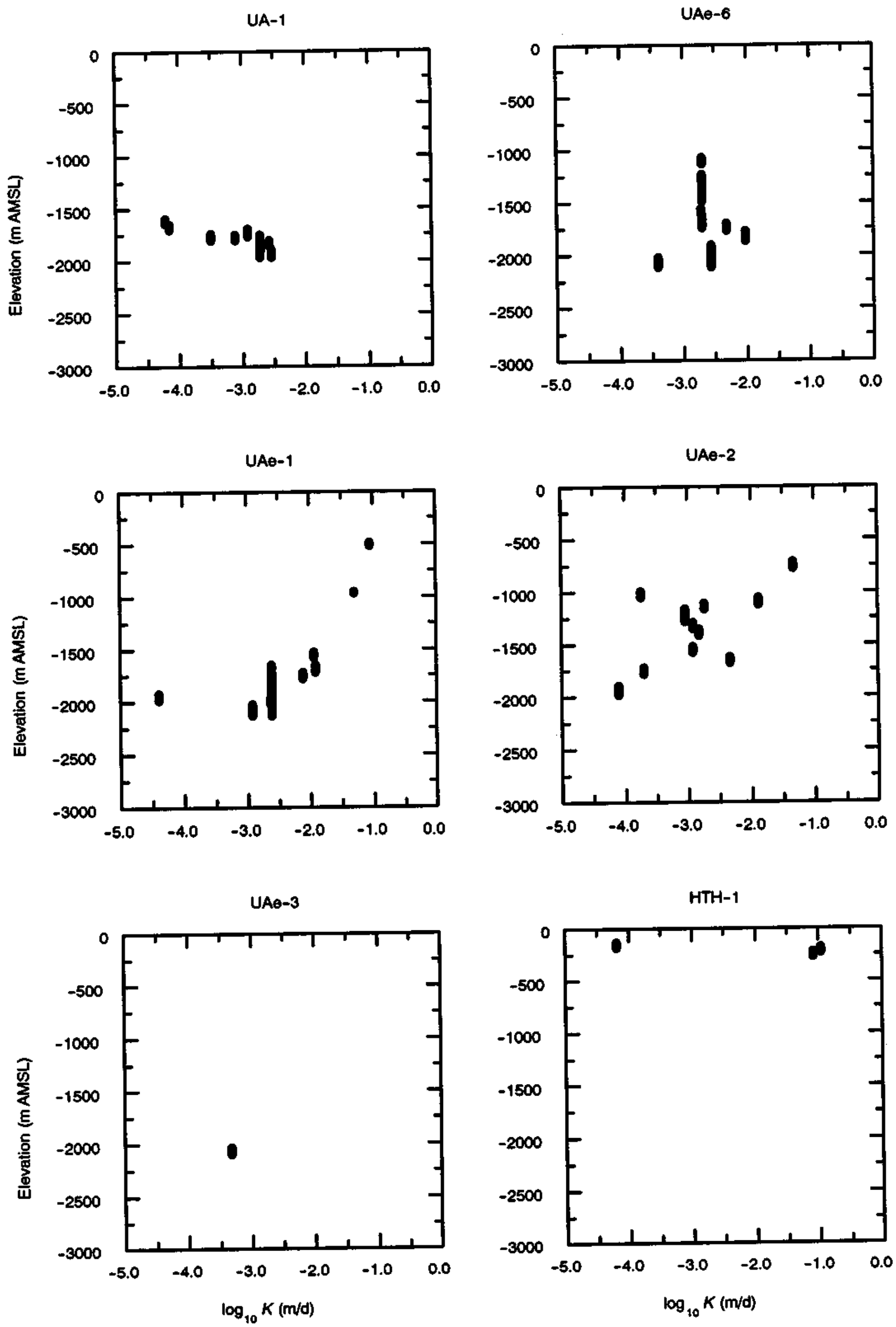

Figure 2.2. Variation with depth of hydraulic conductivity values obtained from packer tests.

recovery after a swabbing test, are used in the flow model calibrations. These water levels are assumed to represent hydraulic head at the depth of their measurement. The influence of groundwater salinity and temperature on the measurements is unknown because these parameters were not consistently reported with the water level data, thus the measured head values could not be corrected to represent equivalent freshwater heads. This uncertainty increases with increased depth as the salinity and the heat effects on water density increase with depth.

Measurements of static water levels in various shallow observation wells are used to augment the packer test data, which were generally collected at depths greater than 500 m. Construction details and exact locations of many of the shallow wells are not available, but water levels consistently indicate that the water table lies within several meters of ground surface and that heads decline with increasing depth (Table 2.2).

Table 2.2. Shallow hydraulic head data from the three testing areas.

Name	Elevation (m)	Depth (m)	Bottom Elevation (m)	Water Depth (m)	Water Elevation (m)
Long Shot Site					
W-1	42.1	1.1	40.9	0.0	42.1
WL-1	42.1	2.7	39.4	0.8	41.3
WL-2	42.1	3.4	38.7	1.0	41.1
Well-1	42.1	37.8	4.3	1.7	40.4
Well-3	42.1	37.5	4.6	2.5	39.6
Well-8	42.1	37.5	4.6	0.8	41.3
Well-9	42.1	100.9	-58.8	4.7	37.4
USGS	42.1	213.3	-171.3	0.6	41.5
Dudley #1	42.1	30.5	11.6		42.0*
Dudley #2	42.1	91.4	-49.3		38.8*
Dudley #3	42.1	121.9	-79.8		38.3*
Dudley #4	42.1	152.4	-110.3		37.3*
Cannikin Site					
White Alice	88.4	96.3	-7.9	1.8	86.6
HTH-3	52.5	51.2	1.3	13.5	39.0
Milrow Site					
W8	29.3	1.6	27.7	0.4	28.9
W10	34.6	2.1	32.5	0.9	33.7
W11	34.0	1.5	32.5	0.6	33.5
W14	37.3	2.1	35.3	1.6	35.7

*estimated from graph

2.2.3 Porosity

Measurements of total porosity of rocks undisturbed by nuclear test effects were made on 197 core samples obtained from UAe-1, UAe-2, UAe-3, and UAe-6c (Lee, 1969a,b,c,d), giving an arithmetic mean of 0.14 and a standard deviation of 0.08 for samples from all depths, and a mean

Breccias (Figure 2.3). These measurements represent the matrix porosity of blocks of basalt and breccia that occupy the volume between connected fractures, and are likely to overestimate the effective flow porosity in the fracture zones themselves. The U.S. Army Corps of Engineers and USGS (1965) estimated an effective porosity of 1.0×10^{-2} from hydraulic testing in EH-1, and 1.0×10^{-3} from hydraulic testing in EH-5 (both at the Long Shot site). Nork *et al.* (1965) assumed a value of 1.0×10^{-3} for flow in joints in andesite and a value of 1.0×10^{-2} for combined fracture and interstitial flow at the Long Shot site. Essington *et al.* (1970) used effective porosities of 1.5×10^{-3} to 2.5×10^{-3}. By analyzing the response in well UAe-1 to tidal and barometric effects, Fenske (1972) estimated a fracture system porosity of 1×10^{-3}. Dudley *et al.* (1977) used estimates of fracture spacing and *K* data and an empirical relationship developed by Snow (1968) to estimate fracture porosities between 1.9×10^{-5} and 6.3×10^{-4}. Though the hydraulic response estimates have the fewest uncertainties, the lower value of 5.0×10^{-4} is selected in the present study to represent the fracture porosity of undisturbed rock at Amchitka to be conservative. This is considered a mean value around which a random distribution is generated and the porosity effects on transport are evaluated.

Direct measurements of porosity are not available for the chimney regions of the Amchitka nuclear tests, but indirect estimates have been made. Using hydraulic properties of the surrounding rock, estimates of chimney infill, chimney dimensions, and measured water level rise, Fenske (1972) calculated the distribution of rubble porosity in the Milrow chimney to be zero (in the puddle glass) at the bottom and 0.14 at the top. Claassen (1978) used many of the same types of data, but a different analytical technique, and estimated the distribution of porosity in the Cannikin chimney to be 0.10 near the bottom and 0.04 near the top. Claassen (1978) provides no explanation for the

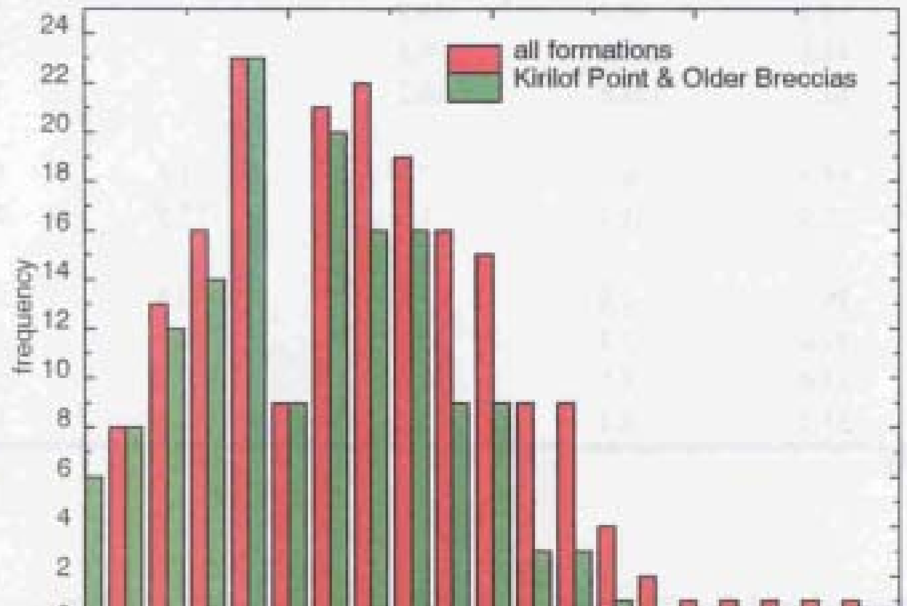

Figure 2.3. Variation of total porosity determined from laboratory tests on core samples.

to be 0.10 near the bottom and 0.04 near the top. Claassen (1978) provides no explanation for the very different distribution of rubble porosity compared to that of Fenske (1972). Garber (1971) provides an estimate of chimney porosity for the Bilby test at the NTS, which was detonated in saturated zeolitized volcanic tuff. By comparing the volume of water removed during a pumping test with the observed interruption in chimney infill, Garber (1971) estimated the porosity to be about 0.07 at a level 110 m above the working point, and about 0.02 at a level 175 m above the working point. A value of 0.07 falls within the range of all these estimates and is selected to represent the rubble porosity for the Milrow chimney.

2.2.4 Recharge

Dudley *et al.* (1977) compared precipitation records and runoff measurements and conclude that most of the precipitation results in surface-water runoff. They note that the water table is at or very near land surface over most of the island, as indicated by the many lakes and streams and water levels in shallow holes. They do note that recharge seems to take place in the months of July and August, following the two months of lower precipitation in May and June (accompanied by water table declines). During this period, there is less correspondence between rainfall and runoff, Dudley *et al.* (1977) note that the tundra, peat, and fractured and weathered volcanic rock of the shallow subsurface are permeable and "where unsaturated, are capable of accepting recharge readily." They also note that this shallow groundwater system behaves locally, discharging in the lakes and streams. They conclude that recharge from precipitation percolates through the mantle of vegetation and colluvium, flowing downslope along the surface of underlying less permeable volcanic rock to issue as springs.

Gonzalez (1977) also evaluated precipitation and runoff and similarly concluded that most of the precipitation results in surface water runoff. Gonzalez observed that the geologic and hydraulic character is generally the same for all three test sites. He described the situation as follows: "A thick cover of turf ranging from a few centimeters to several meters thick is dominant over the entire area. Annual precipitation of over 90 cm collects temporarily in an abundant number of lakes and underlying turf before discharging to streams and finally into the oceans. The runoff that occurs within drainage basins almost simultaneously with rainfalls suggests that the rocks underlying the turf and shallow lakes are either low in permeability, or are saturated to land surface." He further notes that "probably only a small amount of precipitation infiltrates into fractures of deeper rocks." Gages in the watersheds surrounding Milrow and Cannikin accounted for 95 and 93 percent, respectively, of precipitation as runoff. The percentage was lower (57 percent) at Long Shot, attributed to difficulties in gaging because of the high number of lakes. He also notes that surface water base flow is sustained by groundwater discharge from thick surficial materials. He states that the bulk of groundwater flow occurs at shallow depths, discharging as seeps, springs and into lakes and streams.

Fenske (1972) used a hydraulic analysis to calculate the recharge rate necessary to sustain the freshwater lens suggested at UAe-1 from the water chemistry data. He believed the resulting estimate of 8 cm/yr was high, based on observations of stream discharge.

Estimates of groundwater recharge are made here using the temperature profiles measured in several Amchitka Island boreholes. Vertical fluid movement can affect the flux of heat within the earth. Stallman (1960) presented the basic equations for the simultaneous transfer of heat and water within the subsurface and suggested that temperature measurements can provide a means of measuring fluid velocity. Stallman (1965) presented a method for near-surface temperature fluctuations, which assumes the transient flow of both heat and fluid. Bredehoeft and Papadopulos (1965) present the steady-state solution for heat and fluid, which is applicable in deeper systems where temporal heat variations become negligible. The differential equation for steady-state, one-dimensional, simultaneous heat and fluid flow through isotropic and homogeneous porous media is given by:

$$\left(\frac{\partial^2 T}{\partial z^2}\right) - \left(\frac{c_0 \rho_0 V_z}{k}\right)\left(\frac{\partial T}{\partial z}\right) = 0 \tag{2.1}$$

where T is the temperature (°C), z is the vertical Cartesian coordinate (positive downward, cm), c_0 is the specific heat of the fluid (cal/g), ϱ_0 is the density of the fluid (gm/cm^3), k is the thermal conductivity of the solid-fluid complex (cal cm^{-1} s^{-1} °C^{-1}), and V_z (cm/sec) is the vertical component of the fluid velocity (cm/s). Equation (2.1) is strictly applicable in an isotropic homogeneous, fully saturated porous media, as k is a non-linear function of the water content in the vadose zone. Although not all assumptions required by Equation (2.1) are met at Amchitka, the steady-state, low-porosity, and near-saturated conditions provide an approximation to the simultaneous movement of heat and fluid. Bredehoeft and Papadopulos (1965) provide a solution to Equation (2.1) as:

$$\frac{(T_z - T_0)}{(T_L - T_0)} = \frac{\exp\left(\frac{\beta z}{L}\right) - 1}{\exp(\beta) - 1} \tag{2.2}$$

where T_0 is the temperature at the uppermost elevation (°C), T_L is the temperature at the lowermost elevation (°C), T_z is the temperature at vertical location z (cm), L is the total vertical thickness where thermal data are collected (cm), and $\beta = c_0 \varrho_0 V_z L/k$ is a dimensionless parameter that is positive or negative depending on whether V_z is downward or upward. The vertical fluid velocity V_z is determined by non-linear optimization techniques that search for the value of V_z such that there is a minimum difference between the ensemble observed and simulated temperature profiles.

This model is fit to measured temperature profiles in six Amchitka Island boreholes using a value of rock thermal conductivity of 3 x 10^{-3} cal cm^{-1} s^{-1} °C^{-1}, as used by Green (1965) in an analysis of borehole temperatures at the Long Shot site. Measurement error of temperature typically follows a symmetric particle density formation (pdf) (*i.e.* normal distribution with zero mean) and as such does not have a large impact on the uncertainty of the recharge estimate. This is due to the fact that the recharge estimates are essentially derived from the temperature gradient, which implies that the errors in the temperature measurements tend to cancel if large sample sizes are used in the analysis of recharge. The temperature logs were all measured prior to the nuclear tests, but several months after the completion of drilling (with the exception of EH-3), so they should be reasonably representative of natural temperature conditions (Green, 1965; Sass and Moses, 1969). The logs were run primarily in open holes, so there is the possibility of flow in the boreholes, under natural gradients, disturbing the thermal profiles. This flow would be downward, and was noted in some intervals on the logs, and has the effect of causing a potential error for over-predicting the recharge rate using this method.

The model shows good matches to the measured data, particularly in the upper portions of the boreholes (Figure 2.4), using the estimated recharge rates listed in Table 2.3. Note that these estimates are many times lower than recharge rates used by other researchers. For example, Wheatcraft (1995) assumed a value of 10 cm/yr in his models of groundwater flow and mass transport at Amchitka Island. Low recharge rates on the island are consistent with the high rates of surface runoff observed during precipitation events (Dudley *et al.*, 1977). The recharge rate estimated for the Milrow site using the UAe-2 temperature profile is 0.62 cm/yr, Long Shot is 3.39 cm/yr, and Cannikin is 0.45 cm/yr. As discussed in the flow modeling section, recharge is adjusted during calibration at each site, and uncertainty in the parameter is retained throughout the modeling process. The minimum, mean, and maximum recharge values used for each site, based on site-specific calibration, are as follows: Milrow 0.319, 2.078, 7.839 cm/yr; Long Shot 0.809, 3.27, 14.09 cm/yr; and Cannikin 0.809, 3.62, 18.89 cm/yr. The mean values for both Milrow and Cannikin are higher than those suggested by the temperature profiles, but within the range estimated by previous workers.

Table 2.3. Boreholes used for estimates of groundwater recharge.

Borehole	Depth Interval Top (m)	Depth Interval Bottom (m)	Recharge Rate (cm/yr)
UAe-1	0	500	0.45
UAe-2	0	500	0.62
UAe-3	0	500	0.75
UAe-6h	0	500	2.48
EH-1	61	305	3.39
EH-3*	61	259	7.75

*The EH-3 log was run immediately after drilling, with the unstable borehole conditions resulting in greater uncertainty in the representativeness of the log.

An uncertainty analysis was performed on the recharge estimates derived from the thermal gradients, assuming that the errors associated with temperature measurements are symmetrically distributed with a mean of zero. The uncertainty analysis was performed by utilizing the β parameter derived from the individual inversions from each well. To allow for additional uncertainty in the inversion process, ranges of β were calculated assuming various domain lengths (*i.e.*, shallow and deeper temperature profiles). Next, the ratio of β/L was calculated, which effectively normalized the results from different profiles. These ratios were assumed to follow a uniform distribution spanning the entire range (1.87×10^{-4} to 3.34×10^{-3} cm^{-2}) calculated in the inversion process. The

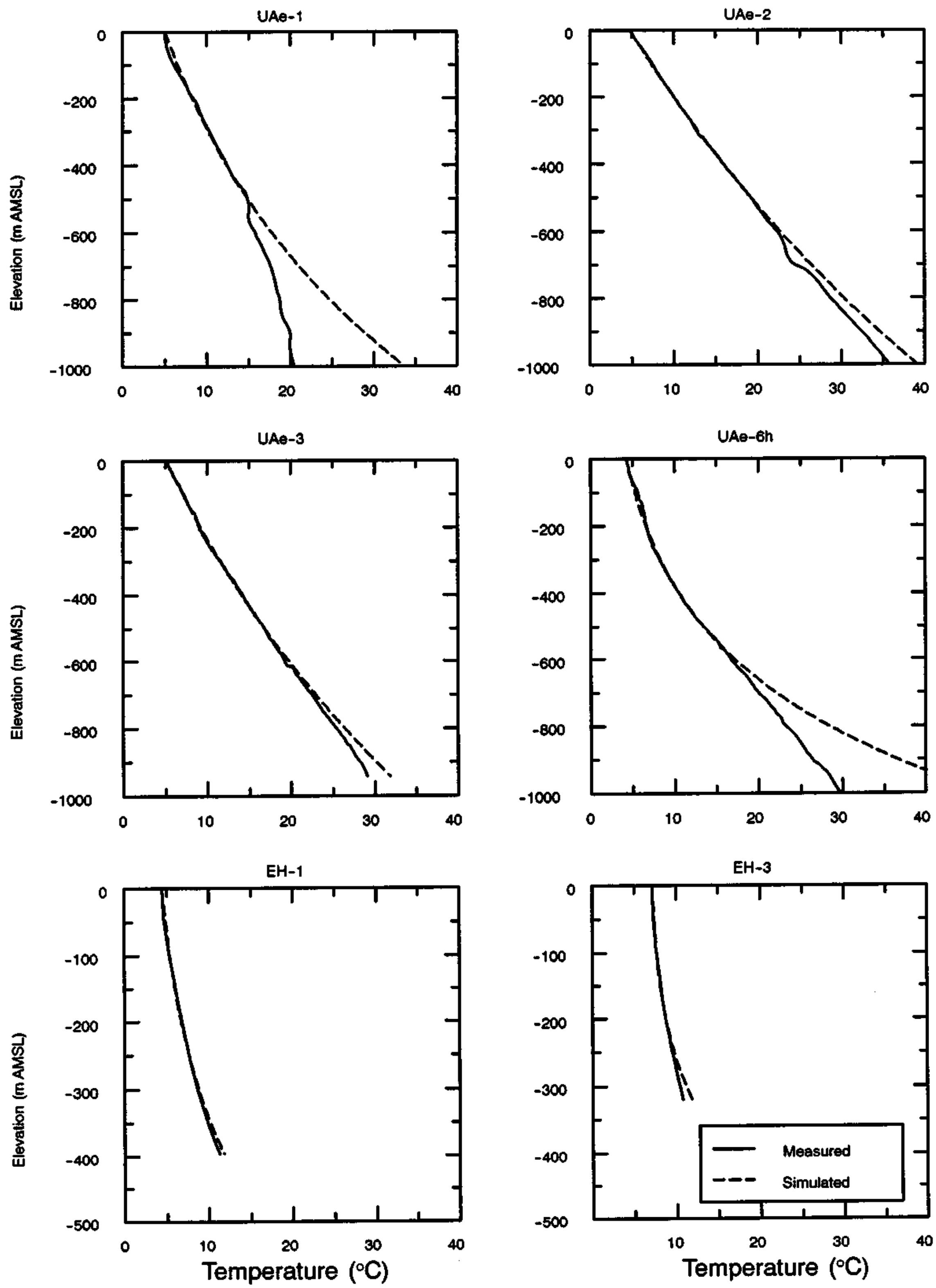

Figure 2.4. Plots of temperature profiles simulated using the groundwater recharge model compared to temperature profiles measured in the boreholes.

uncertainty associated with the thermal conductivity was assumed to be uniformly distributed with a mean equal to 0.003 cal cm^{-1} sec^{-1} °C^{-1} (Green, 1965) and a range of 0.001 to 0.005. It is important to note that the thermal conductivity used in this analysis is of the solid-fluid complex and the use of measured values as determined by Green (1965) is more representative of local conditions than a more generalized basalt conductivity. The relative range used to represent the uncertainty in the thermal conductivity is similar to the relative range suggested by the reviewer. The uncertainty in the thermal conductivity and the β/L parameters were used in a Monte Carlo framework to calculate the expected uncertainty in the recharge rates. One thousand random independent thermal conductivity and β/L values were drawn from their respective distributions and then fluid velocity was calculated according to:

$$V_Z = \frac{\beta}{L}\frac{K}{c_0\rho_0} \tag{2.3}$$

where V_z is the vertical fluid velocity, β is the shape parameter used in the inversion process, L is the length of the profile, K is the thermal conductivity, c_0 is the specific heat of the fluid, and ρ_0 is the density of the fluid. The resulting probability distribution function is shown in Figure 2.5, with a lognormal fit shown simply to note that the distribution has a certain degree of skew which is caused by the multiplication in the V_z calculation. The results suggest that at a 95 percent confidence level, the recharge values should range between 0.3 to 4.8 cm/yr. The mean and minimum values used in the model for the three sites fall within this range, though the maximum end of the model ranges exceeds it.

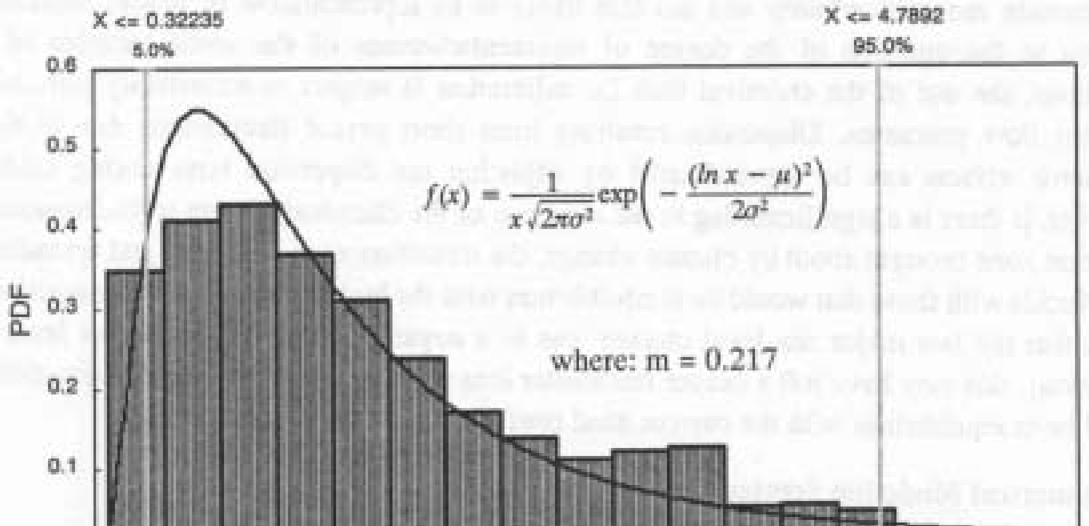

Figure 2.5 A lognormal fit to the recharge values obtained by the uncertainty analysis of the temperature logs.